Effects of Convective Momentum Transport on the Atmospheric Circulation in the Community Atmosphere Model, Version 3

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ABSTRACT

Transport of momentum by convection is an important process affecting global circulation. Owing to the lack of global observations, the quantification of the impact of this process on the tropospheric climate is difficult. Here an implementation of two convective momentum transport parameterizations, presented by Schneider and Lindzen and Gregory et al., in the Community Atmosphere Model, version 3 (CAM3) is presented, and their effect on global climate is examined in detail. An analysis of the tropospheric zonal momentum budget reveals that convective momentum transport affects tropospheric climate mainly through changes to the Coriolis torque. These changes result in improvement of the representation of the Hadley circulation: in December–February, the upward branch of the circulation is weakened in the Northern Hemisphere and strengthened in the Southern Hemisphere, and the lower northerly branch is weakened. In June–August, similar improvements are noted. The inclusion of convective momentum transport in CAM3 reduces many of the model’s biases in the representation of surface winds, as well as in the representation of tropical convection. In an annual mean, the tropical easterly bias, subtropical westerly bias, and the bias in the 60°S jet are improved. Representation of convection is improved along the equatorial belt with decreased precipitation in the Indian Ocean and increased precipitation in the western Pacific. The improvements of the representation of tropospheric climate are greater with the implementation of the Schneider and Lindzen parameterization.

1. Introduction

Convection is one of the most important processes driving tropospheric climate. In addition to distributing heat and moisture and transporting constituents, convection also transports momentum. The significance of convective momentum transport to the global circulation is still questionable as it cannot be directly measured on a global scale. However, vertical transport of horizontal momentum by convection has been estimated to be significant in local observational campaigns (e.g., LeMone 1983; LeMone et al. 1984) and in estimates over larger regions (Holland and Rasmusson 1973; Carr and Bretherton 2001; Lin et al. 2005).

Momentum transport in localized convective events can also be quantified using cloud-resolving models (Soong and Tao 1984; Kershaw and Gregory 1997; Zhang and Wu 2003). These simulations allow for detailed momentum budget calculations within a convective storm, including calculations of in-cloud horizontal velocities that are crucial to the calculation of the vertical momentum flux. In general circulation models (GCMs) convective momentum transport must be parameterized as convection itself is parameterized. The largest uncertainty in parameterizing convective momentum transport in GCMs comes from the need to estimate in-cloud horizontal velocities. In early convective momentum transport parameterizations (e.g., Schneider and Lindzen 1976, hereafter SL76), in-cloud horizontal velocities were assumed to be dependent only on lateral entrainment and detrainment rates; however, observations by LeMone et al. (1984) and all cloud-resolving model studies find that the in-cloud pressure gradients have a large influence on in-cloud horizontal velocities. In recent years, Zhang and Cho (1991), Wu and Yanai (1994), and Gregory et al. (1997,
hereafter GKI97) have developed parameterizations of momentum transport by convection that include the effects of the pressure gradient on in-cloud velocities. These parameterizations all make assumptions about the pressure distribution in a convective storm and, since the tuning factors in these parameterizations are based on a handful of mesoscale model simulations, they still hold a fair degree of uncertainty.

Due to the potential importance of convective momentum transport on the global scale, parameterizations of convective momentum transport have been included in several GCM studies. Using a simple parameterization of convective momentum flux by SL76, Helfand (1979) found an intensification of the upper branch of the Hadley circulation. A similar finding was noted by Zhang and McFarlane (1995), who used the Zhang and Cho (1991) convective momentum transport parameterization, and by GKI97. Both of these studies also noted a reduction of midlatitude westerly jets. On the contrary, Tiedtke (1989) found little influence of convective momentum transport on the Hadley circulation; however, he did note a reduction of 200-mb zonal wind biases over the western Atlantic, South America, and the Pacific.

Recently, Wu et al. (2003), using the Community Climate Model, version 3 (CCM3) with the Zhang and Cho (1991) convective momentum transport parameterization, found large changes in the winter Hadley circulation: a weakening in the northern ascending branch below 700 mb and an enhancement in the midtroposphere in the southern ascending branch. They also found an improvement in the representation of the intertropical convergence zone (ITCZ), primarily through the reduction of precipitation along the equatorial belt (5°S–5°N). Here we present the effects of implementing the SL76 and GKI97 convective momentum transport parameterization in the Community Atmosphere Model, version 3 (CAM3). The goal of this work is to examine in detail how convective momentum transport affects the global circulation. This is accomplished by a detailed analysis of the atmospheric momentum budget and an additional simulation isolating the role of zonal convective momentum transport.

Although the parameterization of SL76 is likely not very realistic due to the lack of treatment of the pressure gradient term, we find the comparison of the SL76 and GKI97 schemes useful as it elucidates what effect the tunable parameters in convective momentum transport parameterizations have on the global circulation.

This paper is organized as follows. Section 2 includes the model description, section 3 discusses the implementation of the two convective momentum transport parameterizations in CAM3, and section 4 contains the results. We end with a discussion and conclusions in section 5.

2. Model description

For the simulations presented here, a modified version of the CAM3 was used. The model was run with the finite volume dynamical core based on Lin (2004) with resolution of 1.9° × 2.5° and 26 levels in the vertical. CAM3 includes the deep convective parameterization of Zhang and McFarlane (1995) and the shallow/middle-tropospheric convective parameterization of Hack (1994). A full description of physical parameterizations can be found in Collins et al. (2006).

The control and modified simulations were run for 20 yr and were forced with climatological sea surface temperatures. We find that the response to the parameterizations is similar even after 10 yr of simulation, and the patterns of changes in the two simulations are similar, suggesting a robust response. We call the CAM3 simulation with the SL76 parameterization “SL” and the simulation with the GKI97 parameterization “GR.” Note that the cumulus parameterizations were applied only to the deep convection scheme of Zhang and McFarlane (1995).

3. Convective momentum transport

a. Formulation

In a general circulation model, convective momentum transport is implemented as an added drag on the mean flow. Since it is a parameterized process, the acceleration of the mean flow owing to convective momentum transport, or “cumulus friction” $F_c$, is calculated in each grid cell containing convection, and the resulting tendency is added to the zonal and meridional momentum equations.

Following Kershaw and Gregory (1997), cumulus friction $F_c$ can be expressed as

$$F_c = -\frac{\partial}{\partial \rho} \vec{v} \omega = \frac{\partial}{\partial \rho} [M_{u}(\vec{v}_u - \bar{\vec{v}}) + M_{d}(\vec{v}_d - \bar{\vec{v}})].$$

In the above, the subscripts $u$ and $d$ are used to depict updraft and downdraft averaged quantities, respectively, and an overbar represents an environmental average. The rightmost component of Eq. (1) represents a traditional approximation used in convective parameterizations. It is strictly valid only when the updraft and downdraft areas constitute a very small fraction of the gridbox area. Here $M$ is the mass flux, $\vec{v}$ is the horizontal velocity vector, and $\omega$ is the vertical pressure velocity. The above formulation neglects the contribu-
tion of the environment to the momentum transport and assumes that intraupdraft correlation between the horizontal and vertical velocities are small (Kershaw and Gregory 1997). In a GCM, the difficulty in calculating cumulus friction comes from the need for estimating the in-cloud horizontal velocities $v_u$ and $v_d$.

The SL76 parameterization assumes that in-cloud velocities are dependent only on lateral entrainment rates, whereas the GKI97 parameterization also accounts for the influence of the pressure gradient on the in-cloud velocities. Pressure gradient terms are introduced into the mass continuity equation for the bulk updrafts and downdrafts, $P_G^u$ and $P_G^d$, respectively. The in-cloud velocities are therefore calculated from the following:

$$-\frac{\partial (M_u v_u)}{\partial p} = E_u \nabla - D_u v_u + P_G^u,$$

$$-\frac{\partial (M_d v_d)}{\partial p} = E_d \nabla + P_G^d,$$

where $P_G^u$ and $P_G^d$ are equal to zero in the SL76 parameterization. The GKI97 parameterization specifies the pressure gradient terms based on empirical relationships:

$$P_G^u = -C_u M_u \frac{\partial \nabla}{\partial p},$$

$$P_G^d = -C_d M_d \frac{\partial \nabla}{\partial p},$$

where $C_u$ and $C_d$ are tunable parameters. In this implementation of the scheme we use $C_u = C_d = 0.7$ following GKI97. The value of $C_u$ and $C_d$ control the strength of convective momentum transport. As these coefficients increase so do the pressure gradient terms, and convective momentum transport decreases. Convective momentum transport is weaker in the GR simulation compared to the SL simulations since the addition of the pressure gradient term in the GR simulation reduces the difference between in-cloud and environmental velocities, hence reducing the vertical momentum transport.

b. Cumulus friction

Figure 1 shows the cumulus friction for the SL and GR simulations averaged over the months December–February (DJF). As convective momentum transport is
related to convective mass flux, cumulus friction extends into the upper troposphere in the tropics where convection is deep and only up to about 500 mb in the extratropics. Westerly (easterly) cumulus friction in CAM3 reaches 2 m s\(^{-1}\) day\(^{-1}\) \((-3.5\) m s\(^{-1}\) day\(^{-1}\)) with the SL76 formulations, and only 0.8 m s\(^{-1}\) day\(^{-1}\) \((-1.4\) m s\(^{-1}\) day\(^{-1}\)) with the GKI97 formulation. In both simulations, maximum cumulus friction occurs in the winter hemisphere with maximum easterly cumulus friction below 900 mb between 0° and 30°N, whereas maximum westerly cumulus friction occurs above that region and below 850 mb between 30° and 60°N.

The meridional force on the mean flow from convective momentum transport is also significant. Southerly (northerly) cumulus friction reaches 2.0 m s\(^{-1}\) day\(^{-1}\) \((-1.4\) m s\(^{-1}\) day\(^{-1}\)) with the SL76 parameterization and 1.2 m s\(^{-1}\) day\(^{-1}\) \((-0.8\) m s\(^{-1}\) day\(^{-1}\)) with the GKI97 parameterization.

During June–August (JJA) convective momentum transport shows similar features as in DJF in the respective summer and winter hemisphere; however, the magnitudes of westerly convective momentum transport are approximately 50% stronger in JJA (not shown).

4. Results

a. Momentum budget and the meridional circulations

To examine the effects of convective momentum transport on the atmosphere’s momentum budget in more detail, we consider all of the terms in the zonal momentum equation. We define the zonal average operator as

\[
[x] = \frac{1}{2\pi} \int x \, d\lambda.
\]

Departures from the zonal average \([x]\) are defined as eddies:

\[
x^* = x - [x].
\]

Zonally averaged primitive equations, with pressure as the vertical coordinate can be written as

\[
\frac{\partial u}{\partial t} = f(u) - \frac{1}{a^2 \cos \phi} \frac{\partial}{\partial \phi} \left( a \cos^2 \phi \frac{\partial}{\partial \phi} \left[ a^2 \cos \phi \frac{\partial \left( (\omega || u) \right)}{\partial \phi} \right] \right) - \frac{1}{a^2 \cos^2 \phi} \frac{\partial}{\partial \phi} \left( a \cos^2 \phi \frac{\partial \left( \nu^* u^* \right)}{\partial \phi} \right) - \frac{\partial}{\partial \phi} \left( \omega^* u^* \right) + \left[ X_{GW} \right] + \left[ F_{cs} \right] + \left[ D_u \right],
\]

where \(X_{GW}\) is the zonally parameterized gravity wave drag, \(F_{cs}\) is the cumulus friction, and \(D_u\) is the vertical diffusion. The first term on the right-hand side of (8) is the Coriolis force; the subsequent two terms represent the divergence of relative angular momentum flux owing to the mean meridional circulation; and the fourth and fifth terms on the rhs of (8) represent the momentum flux divergence due to resolved eddies.

Figure 2 shows the momentum budget terms averaged over DJF for the control simulation. The various momentum budget terms are calculated from 5 years of 3-hourly instantaneous model output interpolated onto a pure pressure coordinate. Figure 2 shows that the atmospheric momentum budget in the lowest layers of the troposphere, below 800 mb, is controlled by the Coriolis force and vertical diffusion. Thus, in the lower branch of the Hadley cell, the easterly trade winds are maintained by the equatorial flow in the lower branch of the Hadley cell against the frictional dissipation. In the upper troposphere, the zonal mean winds are maintained primarily by a balance between the Coriolis torque and the divergence from resolved eddies. Divergence due to the mean meridional circulation and gravity wave drag contribute to this balance; however, these terms are significantly smaller. Thus, in the upper branch of the Hadley circulation, the zonal winds are maintained by the addition of westerly momentum by the poleward branch of the Hadley cell and removal of westerly momentum by eddies.

In the midlatitudes, where the Ferrel cell exists, the momentum balance is maintained by the same dominant terms as in the tropics, however, the sign of the terms is reversed. For reference, in Figs. 3a–d compare the meridional circulation in the control simulation to the National Centers for Environmental Prediction (NCEP) reanalysis. In comparison to NCEP reanalysis, the upward branch of the Hadley circulation in the Northern Hemisphere in DJF is significantly too strong (Fig. 3c). On the other hand, the upper- and lowermost portion of the upward branch of the Hadley circulation in the Southern Hemisphere is too weak. These differences in the Hadley circulation are even more pronounced when compared with European Centre for Medium-Range Weather Forecasts (ECMWF) reanalysis (not shown). The meridional components of the Hadley circulation also differ from observations (Fig. 3d): the equatorward branch of the Hadley circulation is approximately 50% too strong in the Northern and Southern Hemisphere. The southerly flow in the upper levels in the Northern Hemisphere is slightly too strong.

In JJA (not shown) the strongest upward motion in the observed Hadley circulation is centered on 10°N.
Compared to both NCEP and ECMWF reanalyses, this vertical motion is too weak above 500 mb in the control simulation. In addition, the control simulation exhibits strong vertical motion right over the equator, which is not observed. In JJA strong meridional flow is observed in the Hadley circulation in the winter hemisphere only. Compared to observations, the southerly flow is approximately 40% too strong in the winter hemisphere and too weak between the equator and 10°S. Between 10°S and the equator, the northerly flow in the upper branch of the circulation is too strong.

In the presence of convective momentum transport, the momentum budget of the atmosphere is altered. These changes are shown in Fig. 4. The largest changes near the surface occur in the Coriolis term: over 2 m s⁻¹ day⁻¹ near 15°N and −2.4 m s⁻¹ day⁻¹ near 40°N in the SL simulation (Fig. 4a) and 1 m s⁻¹ day⁻¹ in the GR simulation (Fig. 4b). Vertical diffusion changes near the surface are also significant in the tropics, with magnitudes of over 1.75 m s⁻¹ day⁻¹ in the SL simulation (Fig. 4e) and 1.25 m s⁻¹ day⁻¹ in the GR simulation (Fig. 4f). In the upper tropical troposphere, the largest change also comes from the Coriolis term: approximately 1.0 m s⁻¹ day⁻¹ near 15°N and 15°S in the SL simulation. There are also changes in the eddy divergence in this region, reaching 0.75 m s⁻¹ day⁻¹ in the SL simulation (Fig. 4c). Changes in the momentum budget terms in the upper troposphere are very small in the GR simulation. From Fig. 4, it is apparent that the addition of convective momentum transport is balanced by a change in the Coriolis torque, which implies a change in the mean meridional circulation. The change in the Coriolis torque is hence approximately proportional to the change in convective momentum transport between the simulations:

$$f\Delta[v] \approx -\Delta[F_{xx}].$$

This can be seen by comparing Figs. 4a,b with Figs. 1a,c. The above relationship is consistent with the findings of Helfand (1979).

Changes in the momentum budget impose changes on the mean meridional circulations (see Fig. 5). These changes primarily impact the Hadley circulation since convective momentum transport is strongest in the tropics. In DJF, in both the SL and GR simulations, the upward branch of the Hadley circulation is weakened in
the Northern Hemisphere and strengthened in the Southern Hemisphere (Figs. 5a,c). In the upper portion of the branch, the near-equatorial biases in vertical motion relative to the NCEP reanalysis are almost all removed. The lower northerly branch of the Hadley cell is also significantly improved (Figs. 5b,d). The bias relative to the NCEP reanalysis in meridional near-surface wind decreases from \(1.5 \text{ m s}^{-1}\) to \(0.5 \text{ m s}^{-1}\) in the SL simulation and to \(1.0 \text{ m s}^{-1}\) in the GR simulation. The upward southerly branch of the circulation, however, was strengthened between the equator and \(10^\circ\)N and \(10^\circ\)S in the control simulation. In JJA (not shown), the biases in the near-equatorial upward motion are also improved with the inclusion of convective momentum transport. The bias in the southerly low-level flow in the winter hemisphere is improved by about 50% in both SL and GR simulations. Similarly to DJF, there was an increase in the strength of the upper branch of the Hadley circulation near the equator with the addition of cumulus friction.

Up to this point, we have focused our discussion on the atmosphere’s zonal momentum budget. As was shown in Fig. 1, cumulus friction also acts in the meridional direction, primarily below 800 mb. The zonally averaged meridional momentum equation in pressure coordinates can be written as

\[
\frac{\partial \mathbf{v}}{\partial t} = -f \mathbf{u} - \frac{1}{a \cos \phi} \frac{\partial}{\partial \phi} (\cos \phi \mathbf{u} \cdot \mathbf{v}) - \frac{\partial}{\partial p} \left( \omega \mathbf{v} \right) - \frac{1}{a} \frac{\partial}{\partial \phi} \left( \cos \phi \mathbf{u} \cdot \mathbf{v} \mathbf{v} \right) - \frac{\partial}{\partial p} \left( \omega \mathbf{v} \mathbf{v} \right) - \frac{1}{a} \frac{\partial}{\partial \phi} \left( \omega \mathbf{v} \mathbf{v} \right) - \frac{1}{a} \frac{\partial}{\partial \phi} \phi \mathbf{v} \mathbf{v} \mathbf{v} \mathbf{v} \\
+ [Y_{GW}] + [F_{cy}] + [D_v].
\]

In the above \(Y_{GW}\) is the meridional parameterized gravity wave drag, \(F_{cy}\) is the meridional cumulus friction, \(D_v\) is the meridional vertical diffusion, and \(\Phi\) is the geopotential. In the meridional direction, the momentum budget is dominated by the Coriolis term, \(-f \mathbf{u}\), and latitudinal gradient of geopotential (not shown). Where these terms are small (less than \(10 \text{ m s}^{-1} \text{ day}^{-1}\)), near the equator and near the poles, especially in the lowermost atmosphere, the vertical diffusion also be-
comes significant. The meridional cumulus friction can hence directly influence the meridional momentum budget only in the vicinity of the equator. Addition of meridional cumulus friction influences the meridional momentum budget through a balance of the following terms:

\[
\frac{f}{H} \frac{\partial}{\partial \phi} - \frac{\partial D}{\partial \phi}; \quad F_{cy}.
\]

Although meridional cumulus friction can influence the meridional momentum budget, it is clear from (11) that meridional cumulus friction does not directly influence the meridional wind. [In a monthly or seasonal mean, the time change of \([u]\) is very small compared to the other terms in (10).] This is an important point, as previous authors (e.g., Wu et al. 2003) have attributed changes in the Hadley circulation directly to meridional convective momentum transport.

To illustrate further the small influence of meridional momentum transport on the zonally averaged meridional wind, we have repeated the SL simulation with the inclusion of only zonal cumulus friction. We call this simulation “SLU.” Figure 6 shows the difference in meridional wind between the SLU and SL simulations. The figure shows that the meridional wind changes very

**Fig. 4.** Difference of selected zonal momentum budget terms from the control simulation for the (left) SL and (right) GR simulations. Contours are in equal intervals of 0.4 m s\(^{-1}\) day\(^{-1}\), zero contour not shown.
little between these simulations, confirming that it is the zonal convective momentum transport that primarily influences the meridional flow in the tropical circulations.

b. Winds and convection

The inclusion of convective momentum transport in CAM3 strongly influences the near-surface winds as well as the distribution of convection in the tropics. Figure 7 compares the CAM3 surface winds to observations. For observations we use the Quick Scatterometer (QSCAT) satellite scatterometer global wind product (Chin et al. 1998; Milliff et al. 2004). The product is a blend of QSCAT winds averaged for years 2000 through 2004 with the NCEP reanalysis. Comparison of Figs. 7a and 7b shows that in CAM3’s control simulation there is (a) an easterly bias in the tropics in the vicinity of the equator, (b) a westerly bias in the north-eastern Pacific and eastern Atlantic near 20°N, (c) a westerly bias in the North Pacific, and (d) a large westerly bias in the 60°S jet. Although the magnitudes of the surface wind bias appear smaller near the equator as compared to the extratropics, these biases are actually very important because the near-equatorial winds are very weak. Figures 7c,d show that the inclusion of convective momentum transport improves all of CAM3’s biases, in particular in the tropics. The improvements in the equatorial winds reach 2.5 m s⁻¹ in the SL simulation and 1.5 m s⁻¹ in the GR simulation. The extratropical biases are improved, however not completely removed.

In the simulations with convective momentum transport, the zonal winds throughout the troposphere change as a result of the new momentum balance and altered horizontal temperature gradients (not shown). In both DJF and JJA, the tropical winds in the uppermost troposphere (between 100 and 200 mb) increase by a maximum of 2 m s⁻¹ in the GR simulation and by a maximum of 3 m s⁻¹ in the SL simulation (not shown). These changes are lower than those reported by Zhang and McFarlane (1995) and GKI97. In addition, the extratropical westerlies are reduced, but only by 1 m s⁻¹. These changes are limited to the upper half of the troposphere in the SL simulation and extend down to the surface in DJF in the GR simulation. The larger decrease in tropospheric winds in the GR simu-
lution is associated with a stronger increase in temperature throughout the troposphere between 60° and 90°N.

Through changed atmospheric circulation, convective momentum transport also affects the distribution of convection, and hence the surface precipitation rate. This is shown in Fig. 8. For observations we use Global Precipitation Climatology Project (GPCP) data averaged between 1979 and 2003 (Adler et al. 2003). CAM3’s control simulation shows some significant biases in the tropics; in particular, it produced too much precipitation in the Indian Ocean. Smaller biases exist in the Pacific Ocean on and near the equator. Inclusion of convective momentum transport (Figs. 8c and 8d) significantly reduces the precipitation in the equatorial Indian Ocean. In the SL simulation precipitation is reduced by almost 3 mm day$^{-1}$ and in the GR simulation by 1.5 mm day$^{-1}$. In the SL and GR simulations, precipitation is also improved in the western equatorial Pacific.

5. Discussion and conclusions

We have discussed the implementation of two convective momentum transport parameterizations in CAM3. The parameterization of SL76 is a very simple one, which assumes that in-cloud velocities are dependent only on lateral entrainment rates. In addition to lateral mixing, the GKI97 scheme also parameterizes the effects of the in-cloud pressure gradient term on in-cloud horizontal velocities. Since the addition of the in-cloud pressure gradient term reduces the difference between the in-cloud and the environmental horizontal velocities, the effects of convective momentum transport are smaller in the CAM3 simulation with the GKI97 parameterization compared to the SL76 parameterization.

In this paper we examined how convective momentum transport affects the global circulation in CAM3. We have performed a momentum budget analysis to quantify the role of convective momentum transport, especially in the lower tropical troposphere. In addition, to isolate the role of zonal momentum transport we have carried out an additional simulation with the meridional component of convective momentum transport turned off. We find that the zonal momentum transport is primarily responsible for changes in the tropical Hadley circulation, even in the northerly and southerly branches of the circulation. This conclusion
Fig. 7. Annually averaged near-surface winds (m s\(^{-1}\)) for (a) QSCAT observations, (b) Control – observations, (c) SL simulation – Control, and (d) GR simulation – Control. Color shading represents wind speed in (a) and wind speed difference in (b)–(d). The vectors depict wind direction in (a) and vector wind difference in (b)–(d).
Fig. 8. Annually averaged precipitation (mm day$^{-1}$) for (a) GPCP observations, (b) Control – observations, (c) SL simulation – Control, and (d) GR simulation – Control.
differs from that of Wu et al. (2003), who attribute changes in the tropical meridional flow to meridional convective momentum transport.

The zonal component of momentum transport changes the atmospheric momentum budget primarily by changing the Coriolis torque. This is in agreement with the findings of Helfand (1979) and Zhang and McFarlane (1995). Changes in the Coriolis torque imply changes in the tropical Hadley circulation. In both DJF and JJA, convective momentum transport reduces the equatorward flow in the lower winter branch of the Hadley circulation. This significantly reduces the biases in the meridional flow in CAM3 compared to NCEP and ECMWF reanalyses. In this simulation with convective momentum transport, there are also comparable changes in the tropical vertical motion, which improve the representation of the Hadley circulation in CAM3: in DJF, the upward branch of the Hadley circulation is weakened in the Northern Hemisphere and strengthened in the Southern Hemisphere. Convective momentum transport strengthens the southerly (northerly) flow in the upper branch of the Hadley circulation in DJF (JJA) near the equator, slightly increasing the bias relative to the control. Note that the change in the upper branch of the Hadley circulation is much smaller than the changes in the lower branch of the circulation. This differs from the findings of Zhang and McFarlane (1995) and GKI97, who found a significant intensification in the upper branch of the Hadley circulation in the Canadian Climate Center general circulation model and in the Hadley Center Climate Model, respectively. We speculate that this difference is due to the different vertical distribution of convective mass flux and, hence, convective momentum transport in these models.

The inclusion of convective momentum transport affects the distribution of convection and surface winds in CAM3. In the simulations with convective momentum transport, all of the major biases in surface winds, including the tropical easterly bias, subtropical Pacific westerly bias, and bias in the 60°S jet, are improved. Upper-tropospheric tropical winds are accelerated, also improving their representation in CAM3 compared to observations. Convective momentum transport improves some of CAM3’s precipitation biases along the equatorial belt with decreased precipitation in the equatorial Indian Ocean and increased precipitation in the western Pacific.

It was shown in this work that the parameterization of SL76 has a greater effect on the circulation in CAM3 and improves the CAM3 biases more than the formulation of GKI97. As the GKI97 parameterization includes the effects of the in-cloud pressure gradient, it is believed to be more physically representative. However, this parameterization includes tuning parameters: $C_u$ and $C_d$ [see Eqs. (4) and (5)], which may not be set correctly for all convection regimes. In this work, we used $C_u = C_d = 0.7$ following GKI97; however, it was suggested by Zhang and Wu (2003) that these values are too large; hence convective momentum transport might be underestimated in the GR simulation, yielding a worse simulation compared to the SL simulation.

It is also important to note that uncertainty in including convective momentum transport in a global model does not only come from the pressure gradient term. Convective momentum transport is also strongly dependent on convective mass flux [see Eq. (1)], which is itself a parameterized quantity. Any errors in convective mass flux, and the associated entrainment and detrainment rates, owing to an inadequate convective parameterization will therefore carry over into the representation of convective momentum transport. The CAM3 convective parameterization of Zhang and McFarlane (1995) produces mass flux profiles that typically decrease rapidly with height, causing the largest momentum transport in the lowest layers of the atmosphere. This may not be realistic. Hence, although the inclusion of the SL76 parameterization produces a better representation of climate in CAM3 as compared to the GKI97 parameterization, it cannot be determined at this point which parameterization yields a more realistic representation of global convective momentum transport.

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